

THE DEEP ROOTS OF GEOTHERMAL SYSTEMS IN VOLCANIC AREAS: BOUNDARY CONDITIONS AND HEAT SOURCES IN RESERVOIR MODELLING

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ABSTRACT

The energy sources of geothermal systems in volcanic areas are cooling and solidifying intrusions in the crust. The nature of the heat transfer from these hot bodies to the groundwater is not known in detail—it is not known at what depth this interaction takes place, nor is it known at what depth ranges water circulation occurs.

When geothermal modelers construct a model of a production field, this issue is normally avoided. The heat sources are believed to be below the depth range of the model and are accounted for by choosing appropriate boundary conditions in the bottom layer of the model. Another standard practice in geothermal modeling is to drive the model using steady boundary conditions for a long period of time until it reaches a steady state, the so-called natural state, before production is simulated.

This approach to modeling geothermal systems in volcanic areas has worked reasonably well for industrial purposes. However, a model that does not include the entire water circulation is not complete. Recently a well was drilled into magma in the Krafla Geothermal Field in North-Iceland at a depth of 2.1 km (Friðleifsson et al., 2010). This drilling, and an older example of a well in Nesjavellir in the Hengill Area, Southwest Iceland, that was unexpectedly drilled into a very hot formation ($> 380^{\circ}\text{C}$) in 1986 at a depth of 1800 m (Steingrímsson et al., 1990), has shown that the industry cannot overlook this imperfection in geothermal modeling. These examples from Krafla and Nesjavellir, along with other observations, indicate that the roots of geothermal systems in the volcanic active zone in Iceland are at shallower depths than previously believed. Thus, the structure of existing numerical models has to be

reconsidered: Heat sources must be incorporated within the models.

Including heat sources within models also challenges the practice of driving the models into a steady state or natural state, since intrusions in the roots of volcanoes are not steady phenomena. Periods of high intrusion activity are known in volcanic systems in Iceland. In an improved modeling scheme, the heat sources should be included in any model, along with the entire water circulation, and the heat sources should be time dependent.

INTRODUCTION

Numerical modeling of geothermal fluid and heat flow is a valuable tool when operating a geothermal field. It has been used for estimating the production capacity of an operation, its environmental impact, and its sustainability. One of the most commonly used software when constructing and maintaining a numerical model of a geothermal field is the TOUGH2/iTOUGH2 software suite (Pruess et al., 1999). This software has previously been used in simulating the behavior of volcanic geothermal fields under operation in Iceland.

The main operating geothermal power plants in Iceland—Krafla, Nesjavellir, Hellisheiði, Svartsengi, and Reykjanes—are all in active volcanic zones. The energy sources of the systems are cooling and solidifying intrusions in the Earth's crust. How and where the heat transfer between the hot bodies and the groundwater takes place is not known, neither is the depth range of the water circulation. In conceptual models of the geothermal field on which numerical models are based, the heat transfer is assumed to be below the model depth ranges (see, e.g., Björnsson et al., 2003). The deeper roots of the geothermal systems are taken into account by choosing appropriate boundary

conditions for the bottom layer. Heat transfer; i.e., the roots of the geothermal system, is thus avoided. This method has proven to be useful when simulating a geothermal system for practical purposes. A model avoiding the roots of a system is, however, incomplete.

Another useful method when simulating a geothermal production field is to assume that the geothermal system is in equilibrium when production begins. Such a model uses applied boundary conditions and/or steady heat and mass sources until it reaches equilibrium before simulating production. *Equilibrium* in this case means that the system is stable over the period of ~10.000 years. This method is a standard practice in geothermal modeling (O'Sullivan et al., 2001); the main advantage of having a stable system before production is simulated is that all model changes of state are solely due to production. Without artifacts, it is much easier to compare different production scenarios.

The geothermal activity in the volcanic active zones of Iceland is, however, very dynamic making this method questionable. One example of the dynamic nature of geothermal activity is the 1976–1983 rifting period in the Krafla Volcano in Northern Iceland. In a series of events, numerous dykes were formed in the geothermal field of the Krafla Area. Some of these dykes were at relatively shallow depths and did in some cases reach production wells of the Krafla Power Plant. The most dramatic event occurred when a small amount of magma erupted from a well (Einarsson and Brandsdóttir, 1980; Buck et al., 2006).

Another example of the dynamic nature of geothermal systems in Icelandic volcanic zones is the new hot spring area, formed near the village of Hveragerði after an earthquake in May 2008. The Hveragerði village is located in the vicinity of the Hengill Volcano in southwestern Iceland. In this case, higher permeability due to fractures opening was responsible for these changes in the geothermal activity of the area (Hreinsdóttir et al, 2009).

THE ROOTS OF A GEOTHERMAL SYSTEM

Three components are necessary for a geothermal system: a heat source, water, and permeability. These components must be in the correct quantity. Cooling and solidifying intrusions are the energy sources of the geothermal systems in the volcanic zone of Iceland. Water is supplied by precipitation or the ocean, and the movement due to the plate boundary provides the permeability. These components are depicted in Fig. 1. The upper limit of the heat source is d_i , and the water circulation reaches the depth of d_w . Neither of those depths is well known. The heat source is (as said) above a cooling and/or solidifying intrusion. The energy exchange between the intrusion and the groundwater is mainly through thermal conduction, but can also be through degassing of the intrusion.

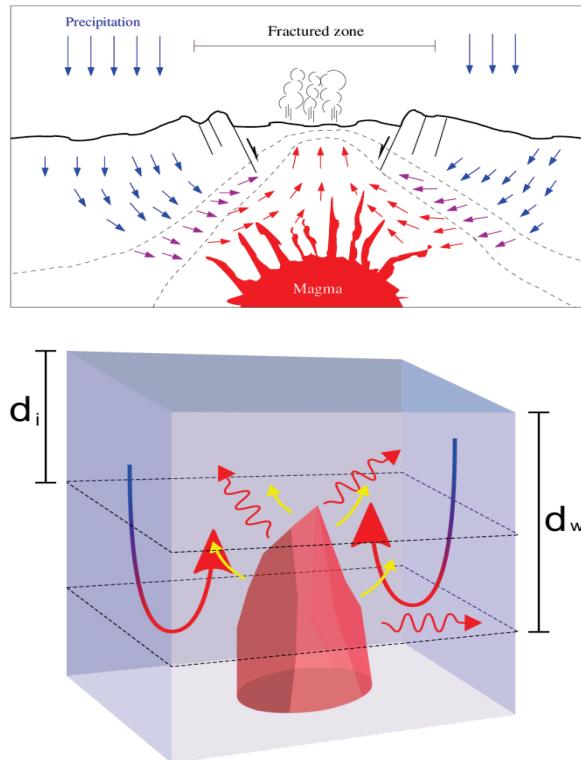


Figure 1. The roots of a geothermal system. The upper part shows the components needed for forming an geothermal system. The lower part shows the depth ranges and the heat transfer. The depth of the source is d_i and the depth range of the ground water circulation is d_w .

The depth of the heat sources are normally taken to be at much greater depths than the depth range of the model. The roots are taken into account by choosing appropriate boundary conditions in the bottom layer of the model—a constant temperature and pressure boundary. The disadvantage of such a constant pressure boundary is that production will increase the inflow of hot fluid into the system significantly. A more conservative method is to inject a fixed amount of fluid and heat into the second deepest layer while having very low permeability in the bottom layer and constant temperature and pressure. Both of these methods are depicted in Fig. 2.

Determining the depth range of the water circulation and the depth of the intrusions driving the geothermal systems is essential. This can partly be done by studying extinct geothermal systems where erosion has opened them up. One has to bear in mind that such a study does not show the system as it was in its most active phase—it shows the entire history of the system. One must also rely on the data that can be sampled using surface geology and geophysical exploration and downhole data (well logs and analysis of cuttings).

A SIMPLE CONCEPTUAL MODEL

Iceland is located at the plate boundary between the N-America and the Eurasian plate. The plates move from each other at an average speed of 2 cm/year. In the conceptual model used here, we assume that the opening space between the plates is filled with 1300°C hot magma. This is known to happen in abrupt events as in Krafla Volcano in the late 1970s–early 1980s. Dykes having thickness on the order of a few meters are formed every one or two centuries. Repeated dyke formation at shallow depths where groundwater is present loads the area with thermal energy and creates a geothermal field.

Icelandic rock, which can be seen on the surface, consist mainly of basalt in the older parts of the country and hyaloclastites in younger areas. Those formations are intersected by few intrusions, mainly dykes. The concentration of intrusions is relatively low except for some specific areas believed to be extinct volcanoes

(see, e.g., Walker, 1963). The concentration of intrusive rocks is also high in wells in geothermal fields. In the Nesjavellir field in the northern part of the Hengill Volcano, the ratio of intrusive rocks is normally above 50% in wells below a depth of 1500 m (Franzson, 1988).

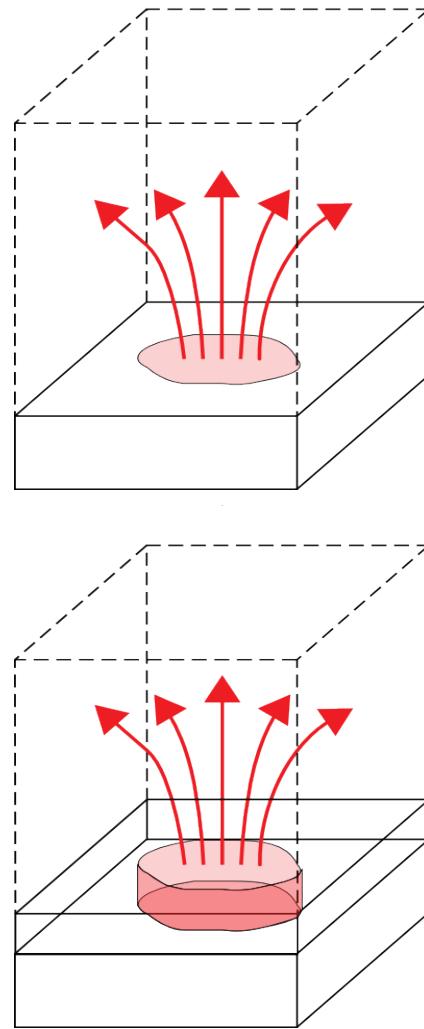


Figure 2. The use of boundary conditions to account for the heat sources of the geothermal systems. The sources are assumed to be below the model's depth range. The upper one shows the constant temperature and pressure boundary and the lower one shows the constant inflow.

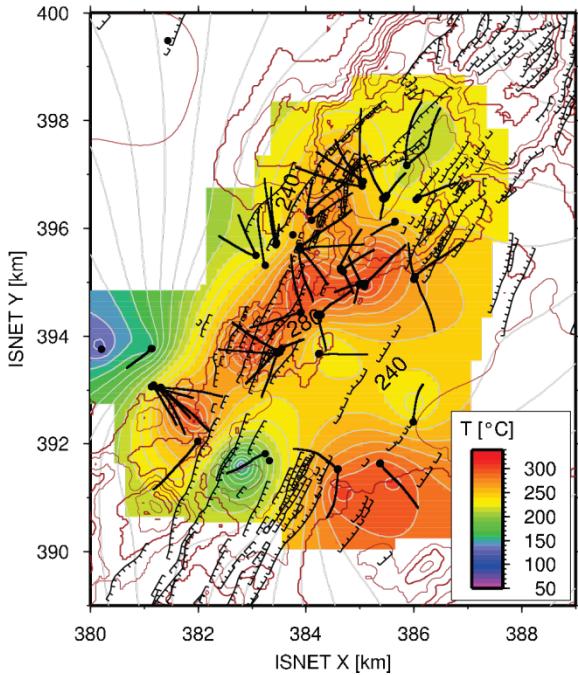


Figure 3. Formation temperature at 1000 m b.s.l in the southern part of the Hengill Area, SW-Iceland.

The abundance of intrusive rocks seen in wells in active geothermal fields supports the hypothesis that dykes and other intrusions at relatively low depths, i.e., within the drilling range, are the fields' energy sources. The depth range of the water circulation is harder to determine. Experience at Krafla, where magma was found at 2.1 km depth (Friðleifsson et.al 2010), and the well data from the Hengill-Area indicate that the water circulation does not reach much deeper than 2–3 km. In the conceptual model used in this study, it is assumed that we have an impermeable bottom at ~2.3 km depth. One can, however, expect that the depth range of the water circulation differs from one field to another in this volcanic zone.

NUMERICAL CALCULATIONS

To test the conceptual model described in the previous section a numerical dummy model was set up. The dummy model is a simplified geothermal field typical of Icelandic volcanic zones. A formation of high permeability housing the geothermal system is surrounded by formations of lower permeability and covered with a cap rock of low permeability. The cap rock separates the geothermal system from the groundwater system above it. It is common to

have different water levels in the groundwater system than in the geothermal system. In the Hellisheiði Field, the typical groundwater level of the geothermal system is ~200 m below surface, but the groundwater level of the upper groundwater system is ~20 m below the surface. The core of the system has a slightly higher permeability than the rest of the geothermal system. The heat sources are mainly placed in the core.

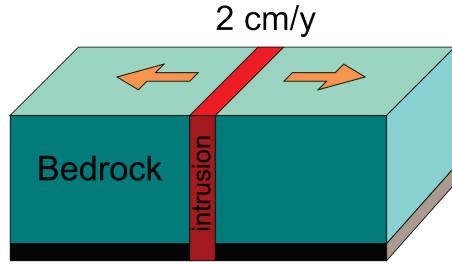


Figure 4. Boundary of two plates moving from each other. The average speed is 2cm/year and magma fills up the gap between the plates forming dykes.

The structure of the numerical model is shown in Fig.5. The energy sources were predefined "dykes" formed every 100 years on an average. The formation of dykes is hard to model in detail. In this work it was assumed that the dykes released their thermal energy into the elements where the dyke was formed. The energy release in each element where the dyke was formed was assumed to obey the relation

$$Q = \frac{E_0}{\tau} e^{-\frac{t}{\tau}} \quad (1)$$

where Q is the energy flow into the element from the dyke, $\tau = 25$ years is the time constant and E_0 is the total energy stored in the magma, i.e., the thermal energy released when a 2 m thick layer of magma within the element cools from 1300°C to the solidification temperature of 1100°C, at which time it solidifies and cools down to the temperature of the geothermal system (here taken as 250°C). There were 14 predefined dykes, i.e., walls of elements from north to south sometimes having small shifts in east-west position. These walls were mainly in

the core of the system and in the four model layers from the depth of 650 m down to 2150 m (see Fig.5). Each of these predefined “dykes” had its own possibility of being formed, but at an average, a dyke was formed every 100 years. Forming of dykes was accounted for by releasing the energy as described in Eq. (1) into the elements of the wall representing the dyke.

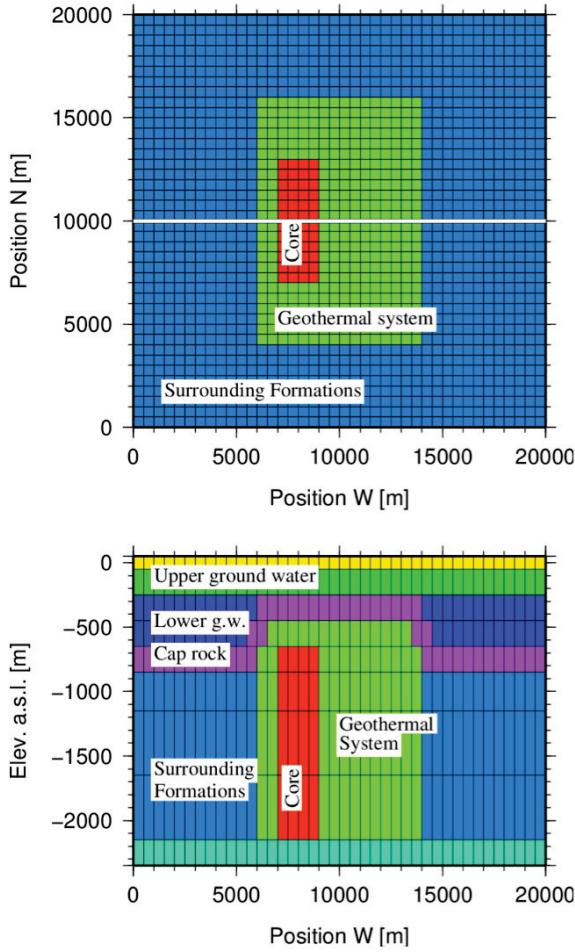


Figure 5. Schematics of the numerical model presented in this work. The upper part shows the model at the depth of 1000 m. The lower part shows the W-E section (marked by the white line the upper part) through the center of the model.

The top layer and the bottom layer of the system had a fixed temperature and pressure, as did the bottom layer. The bottom layer is much thicker, as depicted in Fig. 5, to minimize the effects of the boundary bottom boundary condition, and it was impermeable. In the beginning, the temperature gradient was 100°C/km in the entire

model, and the pressure was in hydrostatic equilibrium. The system was driven using randomly formed dykes, as described above, for 100,000 years. That is a significant time span for the lifetime of a volcanic system in the Icelandic rift zone. It is also a significant time span for a TOUGH2 simulation, because in this case 100,000 years means 1,000 intrusion events, each of them necessarily described in the GENER-block of the input file. The total length of the GENER-block was 278,760 lines when it had been processed, to minimize the number of values describing heat inflow.

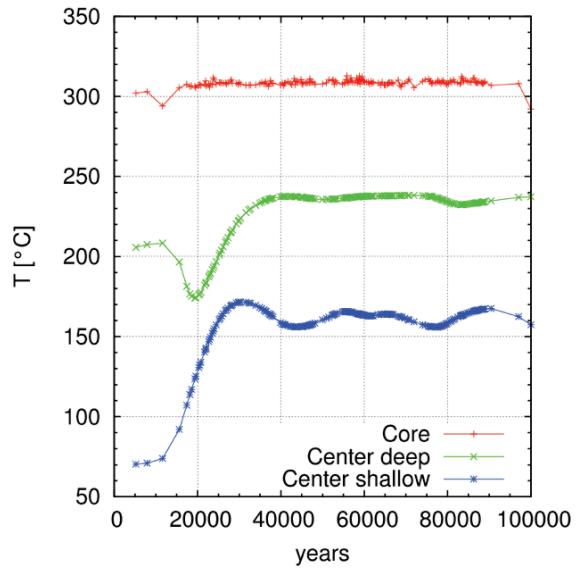


Figure 6. The evolution of temperature at three different locations in the model: In the Core at the depth of 750 m, and at two depths in the center of the geothermal system at of 1850 m (deep) and at 750 m (shallow).

In Fig. 6, the evolution of temperature at three different locations is shown. As can be seen, it takes the system around 30,000 years to reach a metastable equilibrium. The core of the system where the intrusion activity takes place heats up very rapidly, i.e., within the first 2000 years. The outer parts of the geothermal systems take longer to reach equilibrium. In Fig. 7, the temperature at 750 m depth is depicted, as is the temperature in a cross section through the center of the system from west to east. This is the temperature after running the model for 100,000 years. The structure of the temperature is relatively sharp and characterized by a reverse

temperature gradient in the hottest part of the system.

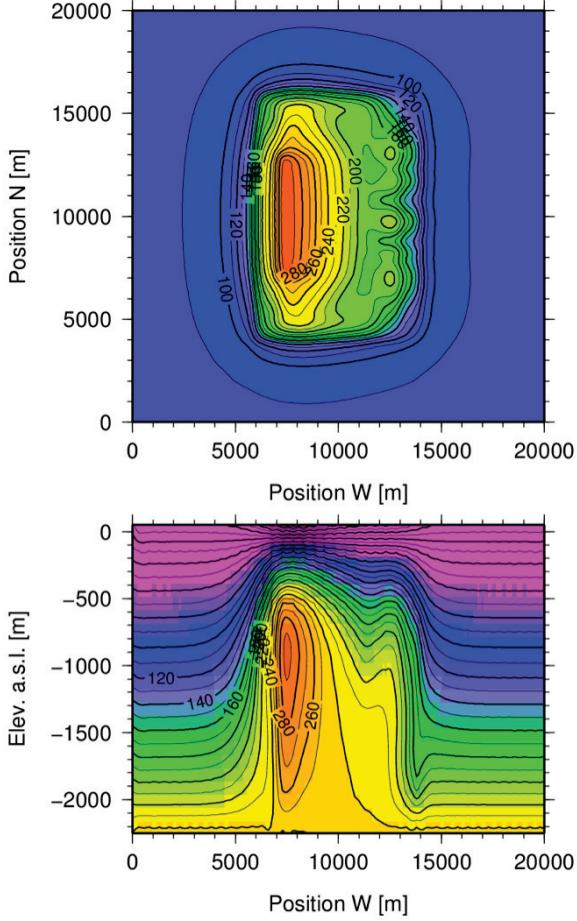


Figure 7. Temperature distribution at the depth of 750 m (above) and in a cross section through the center of the model from west to east (below) as it is calculated by the model. (Same sections as in Fig.5.)

SUMMARY AND CONCLUSION

The standard procedure of building a model for a geothermal system does in many ways avoid the roots of the geothermal system. The energy sources are often accounted for by assuming that they are below the depth range of the model and by choosing appropriate boundary conditions in the bottom layer. This has worked well for practical purposes, as does the assumption that the energy sources are constant in time. However, a model not including the roots of the geothermal system will always be incomplete, and recent cases in which wells were drilled into the heat source of a geothermal system have

shown that it has also practical to incorporate the roots into the numerical models.

In this study, a simple conceptual model of a volcanic geothermal system was used in which it assumed that repeated formations of dykes are the driving mechanism of the system. Implementing the conceptual model using the TOUGH2 code (or any other code) can be carried out by making some simplifications to the problem. The dyke formation was simulated by injecting heat into predefined walls of elements, and a 2 m thick dyke was assumed to be formed (on average) every 100 years. (Note that the expansion of the crust due to rifting cannot be simulated using the TOUGH2 code.) The thickness of the dyke is just an indicator of the amount of energy injected into the predefined walls of elements when dyke formation is simulated. The bottom layer of the model is impermeable, so the entire water circulation is simulated.

The model was run for 100,000 years, with the time dependent heat sources driving it. The model reaches a metastable equilibrium within that time length. When the system has been charged up with heat, it is relatively stable, and temperature fluctuations are relatively small. These features are, however, dependent on model parameters, especially permeability. (The effects of different values for permeability have to be studied further; the time length of each run of the model is strongly dependent on permeability values.) It is, however, astonishing how stable the temperatures are after reaching the metastable equilibrium.

The lateral size of the elements of the model used in this study is relatively big— 500×500 m. The area in which thermal energy injection due to formations of dykes was simulated was much wider than measured in the geothermal field in the Hengill Area. Formation temperature there has relatively narrow structures (See Fig. 3). There is still a lot of work to be done in trying to reproduce formation temperature structures seen in real geothermal systems. For that purpose, the elements must be smaller and the heat injection more concentrated.

We attempted to construct a complete model of a volcanic geothermal system in this work. The entire water circulation was included in the model, as were the heat sources. The question remains open as to whether we will have enough data for including events such as rifting and formation of intrusions as heat sources for our geothermal models.

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